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The Apollo Passive Seismic Experiment

The first lunar seismic experiment is described.

G. Latham, M. Ewing, F. Press, G. Sutton

On or about 20 July of this year, the study of the moon will enter a new phase with the first manned landing on the lunar surface as part of the Apollo 11 mission. In addition to collecting rock samples for return to the earth, the astronauts will install a number of experiments on the lunar surface. This paper describes one of these experiments: the Passive Seismic Experiment. The term "passive" is used because the principal objective of the experiment is to detect naturally occurring seismic events on the surface of the moon. In this sense, it is to be distinguished from the "active" seismic experiments planned for later missions which will include small explosive sources and an array of geophone detectors (*1*).

Because of the very significant contributions to the study of the earth made by seismic methods, seismic experiments have been given high priority for early Apollo missions.

Brief History of the Project

Scientists began to think seriously of the possibility of measuring seismic activity on the moon in 1959 when the Ranger program, designed to land an instrumentation package on the lunar surface, was first discussed. Scientists and engineers at Columbia University's Lamont-Doherty Geological Observatory and the California Institute of Technology's Seismological Laboratory collaborated to produce the first lunar seismometer. The Caltech group took the lead in developing the instrumentation for this program. The Ranger seismometer consisted of a single-axis (vertical component) seismometer with a natural period of 1 second and a total weight of 3.6 kilograms. The seismometer was required to withstand an impact load of 3000*g* (*2*). This seismometer was chosen to fly on three Ranger missions. Unfortunately, one of these missions missed the moon entirely and the other two missions ended in destructive impact.

The next opportunity for a lunar seismic experiment was provided by the Surveyor Soft Landing lunar mission. Development of a seismometer system for the Surveyor mission began at Lamont-Doherty Geological Observatory in 1962. Initial weight limitations and the environmental conditions of the mission were not as severe as for Ranger, and the team of scientists decided to attempt to make a three-axis seismometer system weighing not more than 11.5 kilograms and with sufficient sensitivity at low frequencies to permit measurement of surface waves from moonquakes, if extant, free oscillations of the moon, if excited, and lunar tides. These goals were achieved (*3*) but the seismometer was never flown as part of a Surveyor mission for two reasons. First, decreased payload capability required reversion to a lighter-weight, single-axis seismometer (*4*). Furthermore, mission planners laid greater stress on experiments that would provide information directly applicable to the problems of landing a man on the lunar surface.

Work on the Surveyor triaxial seismic system was reinitiated at Lamont-Doherty in 1966 with the decision by NASA and the scientific community to include a seismometer among the complement of experiments to be placed on the lunar surface by the astronauts of the Apollo program. Now, after 10 years of work and planning, it appears that at last we are nearing the goal of direct measurement of seismic activity on the moon.

Instrument Description

The integrated set of experiments and supporting subsystems which will be installed on the lunar surface by the Apollo astronauts is called ALSEP (Apollo Lunar Surface Experiments Package). The ALSEP central station provides power and telemetry for each of the experiments. Power is supplied by a radioisotope thermal generator.

In order to reduce the astronaut's activities on the first landing (Apollo 11), a simplified version of ALSEP, called EASEP (Early Apollo Scientific Experiment Package), has been constructed for this mission. It weighs 48 kilograms and uses solar cell panels to supply power instead of a nuclear battery. The EASEP package is shown schematically in Fig. 1. The two radioisotope heaters shown are fueled with a small amount of plutonium-238 and produce 15 watts each. These heaters will protect the instrument from the extreme cold of the lunar night. The EASEP system shown will contain only the seismic experiment and is designed to transmit data back to the earth for a nominal period of 1 year (2 years

Dr. Latham is a research associate at the Lamont-Doherty Geological Observatory of Columbia University, Palisades, New York 10964. Dr. Ewing is director of the Lamont-Doherty Geological Observatory. Dr. Press is chairman of the Department of Geology and Geophysics, Massachusetts Institute of Technology, Cambridge 02139. Dr. Sutton is professor at the Institute of Geophysics, University of Hawaii, Honolulu.

Table 1. Scientific measurements of the passive seismic experiment.

Data channel	Parameter measured	Minimum detectable signal	Period	Dynamic range (db)
LPZ	Vertical ground motion	0.3 $m\mu$, peak-to-peak	<15 seconds	90
LPX	Horizontal ground motion	0.3 $m\mu$, peak-to-peak	<15 seconds	90
LPY	Horizontal ground motion	0.3 $m\mu$, peak-to-peak	<15 seconds	90
SPZ	Vertical ground motion	0.1 $m\mu$, peak-to-peak	0.1 second	90
Z-FB	Vertical gravity	8 μgal	∞	60
X-FB	Surface tilt	0.01 second of arc	∞	60
Y-FB	Surface tilt	0.01 second of arc	∞	60
TEMP	Instrument temperature	0.02°C	∞	60

maximum). Operation will be limited to the lunar day since the solar panels must be illuminated to provide power. It is expected that the complete ALSEP system containing at least three additional experiments (for measurements of solar wind and magnetic field) and capable of day and night operation will be included on Apollo 12.

A seismometer consists simply of a mass free to move in one direction suspended by means of a spring, or a combination of springs and hinges, from a framework. The suspended mass is provided with damping to suppress vibration at the natural frequency of the system. The framework rests on the surface whose motions are to be studied and moves with the surface. The suspended mass tends to remain fixed in space because of its own inertia while the frame moves around it. The resulting relative motion between the mass and the frame can be recorded and used to calculate original ground motion if the instrumental constants are known. Conventional seismic instruments are prohibitively large and delicate for use in the lunar mission. A typical single-axis, low-frequency seismometer designed for use on the earth, for example, weighs approximately 30 kilograms and occupies 1×10^5 cubic centimeters. The lunar instrument, containing four seismometers, weighs 11.5 kilograms with a total volume of 2×10^4 cubic centimeters (including a thermal shield).

The Apollo seismometer system consists of two main subsystems: the sensor and the electronics module. The sensor, shown schematically in Fig. 2, contains, in the upper section, three matched low-frequency seismometers with resonant frequencies of 1/15 hertz, aligned orthogonally so as to make possible measurement of the three components of the ground displacement vector. A single-axis seismometer with resonant frequency of 1 hertz is located in the lower section. This seismometer is sensitive only to vertical motion. The "steady" or inertial masses are 750 grams for the low-frequency seismometers and 500 grams for the high-frequency seismometer.

The instrument, including the electronics module, thermal insulation, and a small stool upon which the sensor rests on the lunar surface, is constructed principally of beryllium and weighs 11.5 kilograms. The sensor without insulation is 23 centimeters in diameter and 29 centimeters high. Total power drain varies between 4.3 and 7.4 watts. Performance specifications are given in Table 1. Nominal instrument response curves are shown in Fig. 3.

Low-frequency, horizontal-component seismometers are very sensitive to tilt and must be leveled to high accuracy. In the Apollo system, the seismometers are leveled by means of a two-axis, motor-driven gimbal to within 2 seconds of arc from the local vertical direction. A third motor adjusts the low-frequency, vertical-component seismometer in the vertical direction. Motor operation is controlled by command. These elements are shown schematically in Fig. 4; the low-frequency seismometers are mounted in crisscross fashion to achieve a minimum volume configuration.

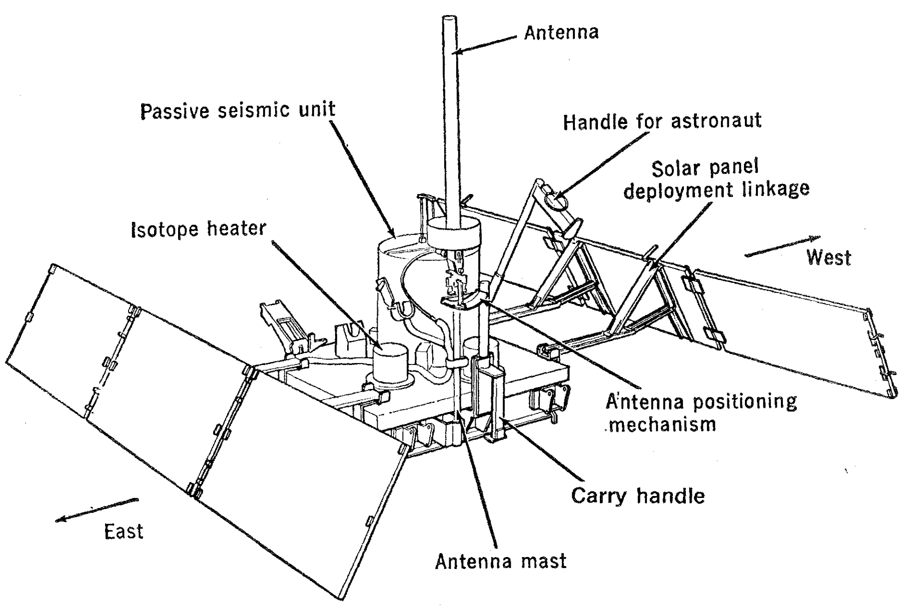


Fig. 1. Passive seismic experiment package (EASEP) as it will appear erected on the lunar surface. [Courtesy of the Bendix Corporation]

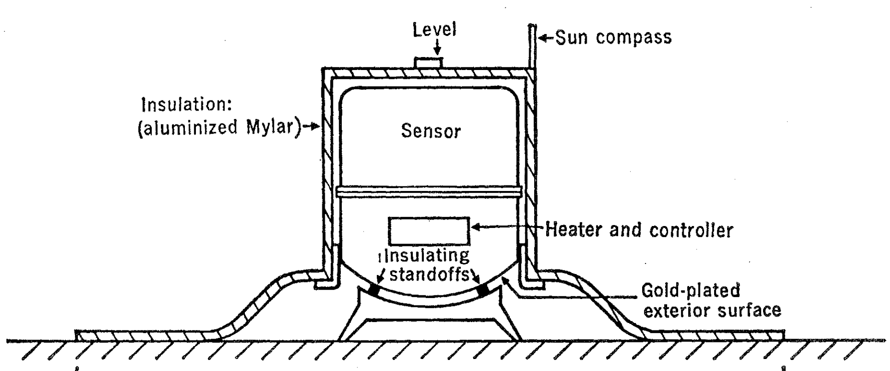


Fig. 2. Schematic diagram of the Apollo passive seismic system.

Low-frequency seismometers are sensitive to temperature variations. These effects are partially compensated by appropriate mechanical design, but good thermal control is needed. In the Apollo seismometer the thermal control system includes active elements—a heater with a maximum output of 2.5 watts, a temperature sensor, and a proportional controller—and passive elements consisting of insulating wrapping of aluminized Mylar. The system used here is designed to provide temperature control of the local lunar surface. In this way, it is planned to reduce thermally induced tilts of the local surface to acceptable levels. Except for this requirement, much simpler control systems could be employed. The control which has been achieved is approximately ± 1°C for most lunar surface conditions. A separate metal stool will be used in all missions after EASEP. The stool will be planted firmly on the lunar surface by the astronaut and will serve as a platform upon which the main instrument housing will rest. The stool provides electrical and thermal isolation from the lunar surface by means of epoxy-glass pads located at the contact points between the sensor base and the stool.

Leveling and azimuth indicators are located at the top of the sensor (Fig. 2). By means of these indicators, the astronaut will level the seismometer and will read its azimuthal orientation to within 5°. The level sensor is simply a small ball in a spherical cup, and the azimuth is read from the shadow cast on the flat top surface of the instrument by the vertical rod.

The instrument is controlled from the earth by a set of 15 commands which control such functions as speed and direction of leveling motors, instrument gain, and calibration. The complete instrument, with the thermal insulation removed, is shown in Fig. 5. The outer surface is gold-plated to improve thermal control.

The electronics module contains variable-gain amplifiers and filters for each seismometer, command logic circuits, motor drive circuits, a power converter, a multiplexer, and an analog-to-digital converter. The output of digital data from the electronics module is combined with engineering data within the central station and is telemetered back to the earth at 1060 bits per second. Figure 6 shows the instrument with its thermal insulation spread out to its full diameter of 1.5 meters as intended for lunar operation.

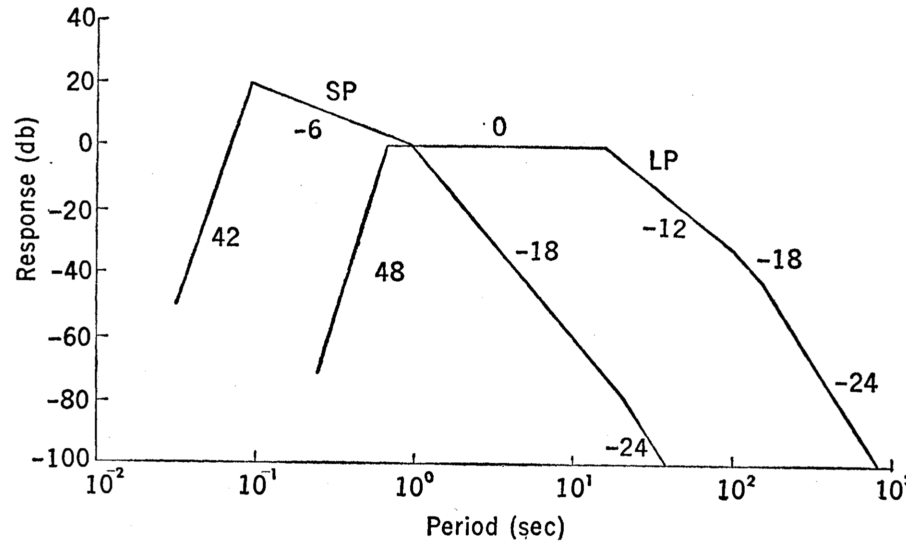


Fig. 3. Nominal response curves of the seismometers; *SP* denotes the short-period seismometer, and *LP* denotes the long-period seismometers. The 0-db level corresponds to a sensitivity of 15 volts per micron of surface motion (maximum gain position).

A major problem in any low-frequency elastic suspension is long-term drift. In the Apollo seismic system, instrumental drift is reduced by feedback as described by Sutton and Latham (*3*). The operating principle of a feedback-controlled seismometer (vertical component) is illustrated in Fig. 7. A voltage proportional to the displacement of the seismic mass from its center position is generated by means of the differential capacitor plate assembly and associated circuitry. This signal is amplified, filtered (high-pass), and amplified again to give the direct seismic output. The high-pass output filter eliminates the effects of initial centering error and long-term drift on the seismic output.

Part of the output signal is fed back degeneratively, through a low-pass filter, to the coil and magnet assembly provided for damping the seismometer. The resulting force on the seismic mass acts to return the boom toward its electrical center position, that is, the position at which the transducer output is zero. The effect of feedback is thus equivalent to an additional spring which acts only at long periods. The equivalent spring constant depends upon the feedback loop gain. The period at which feedback begins to affect instrument response depends upon the feedback filter time constant and loop gain. For long-period boom motion the feedback-controlled seismometer behaves exactly as if it were a short-period pendulum; this accounts for the increased long-period stability. All

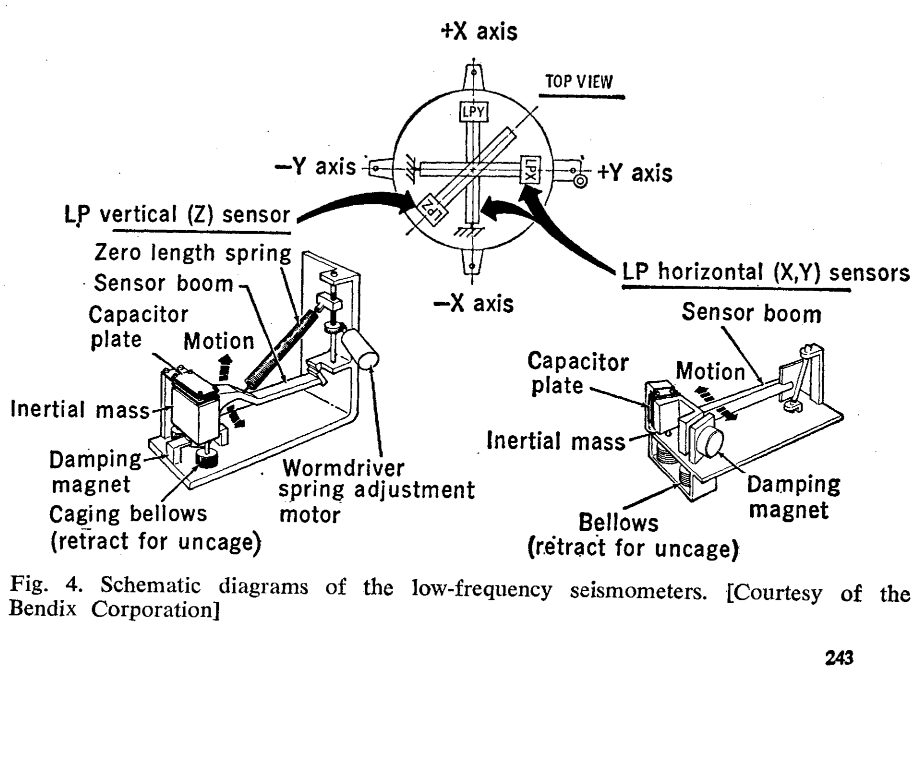


Fig. 4. Schematic diagrams of the low-frequency seismometers. [Courtesy of the Bendix Corporation]

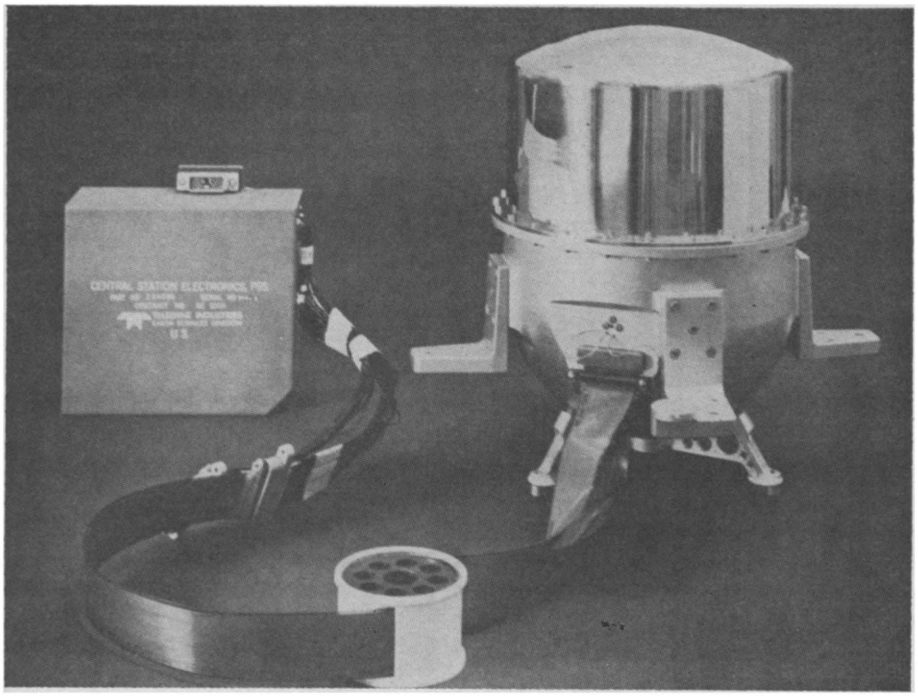


Fig. 5. Photograph of the seismometer system without thermal insulation.

three low-frequency components are feedback-stabilized.

A signal proportional to long-period boom motion is available at the point labeled "feedback output" in Fig. 7. Hence, tidal tilts (on the horizontal components) and changes in the magnitude of gravity (on the vertical component) can be measured by monitoring this signal, provided that long-term disturbances resulting from other causes are sufficiently small.

The short-period seismometer is a velocity-sensitive instrument with a suspended magnet serving as the inertial mass. Relative motion between the magnet and an output coil fixed to the frame generates an electrical signal proportional to the relative velocity. This electrical signal is then amplified and filtered. The suspension system utilizes a cantilever spring as the main support element and "delta rods" in a triangular configuration to keep the moving mass centered. A photograph of the breadboard model is shown in Fig. 8.

A means is provided for the application of an accurate increment or step of current to the coil of each of the four components on command. The current step is equivalent to a step of ground acceleration and will provide a complete calibration of the system.

A caging system must be provided which will secure all critical elements of the instrument against damage during the transport and deployment phases of the Apollo mission. A pneumatic system is used in the present design in which pressurized bellows expand to clamp fragile parts in place. The mechanism is shown schematically in Fig. 4. Uncaging is accomplished by piercing the connecting line on command by means of a small explosive device.

In the Apollo 11 mission, the astronaut will have several tasks to perform for the seismic experiment. After removing the instrument from the lunar module, he will locate the smoothest area within a zone of from 20 to 30 meters from the landing craft. He will place the instrument on the surface, unfold the solar panels, adjust the level of the package to within 5°, orient it in azimuth for maximum illumination of the solar panels, and aim the antenna toward the earth. Immediately upon completion of these tasks, commands will be sent from the Manned Spacecraft Center in Houston to uncage and level the seismometers and to select the proper gain.

If time permits, the astronaut will be asked to stamp on the surface several times to produce observable seismic signals ("astroseisms"). Detectable seismic signals will also be generated by the exhaust pressure of the lunar module during ascent.

Sample records obtained from the Apollo seismometers during functional tests are shown in Figs. 9 and 10. In Fig. 9 the record from an earthquake is compared with the record produced by a conventional seismometer. Such comparisons have shown the Apollo seismometers to be equal or superior in performance to their conventional counterparts in every respect.

Sample records from the feedback outputs showing solid earth tides are presented in Fig. 10. Semidiurnal variations in gravity at the surface of the earth are approximately 0.3 milligal peak-to-peak, and associated tidal tilts are 0.02 to 0.04 second of arc. On the moon, variations in tidal gravity are expected to be approximately 4 times larger, that is, of the order of 1 milligal, and tidal tilts may be 20 to 30 times larger than on the earth (approximately 1 second of arc).

Seismic records, such as those shown

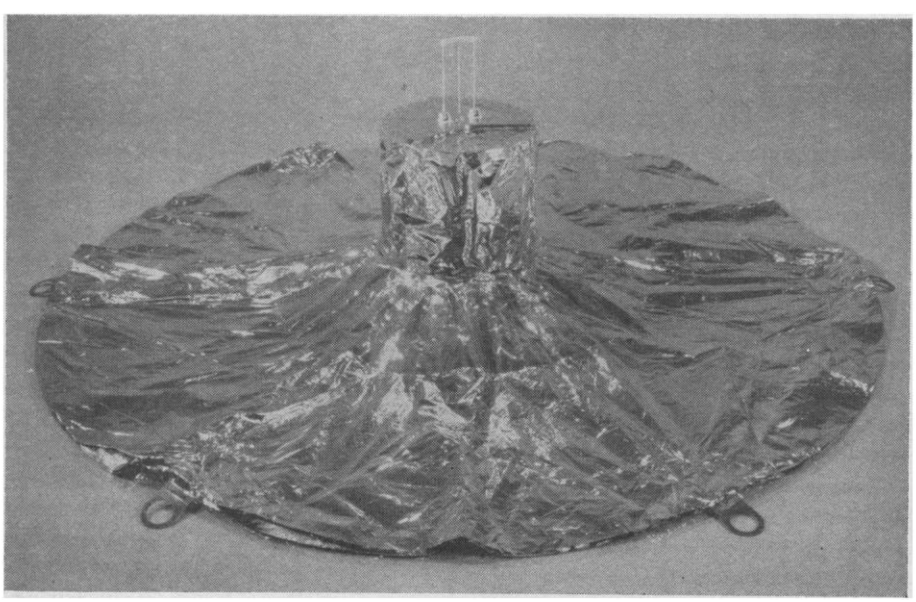


Fig. 6. Photograph of the sensor with the thermal insulation spread out. [Courtesy of the Bendix Corporation]

in Fig. 9, are obtained in vaults with good coupling to basement rock. In early missions, the instrument will very likely be placed on top of loosely consolidated fragmental material which appears to blanket the mare regions with an average thickness of several meters. This will probably result in some attenuation of seismic signal owing to poor coupling with the surface and high absorptivity of elastic wave energy within the material.

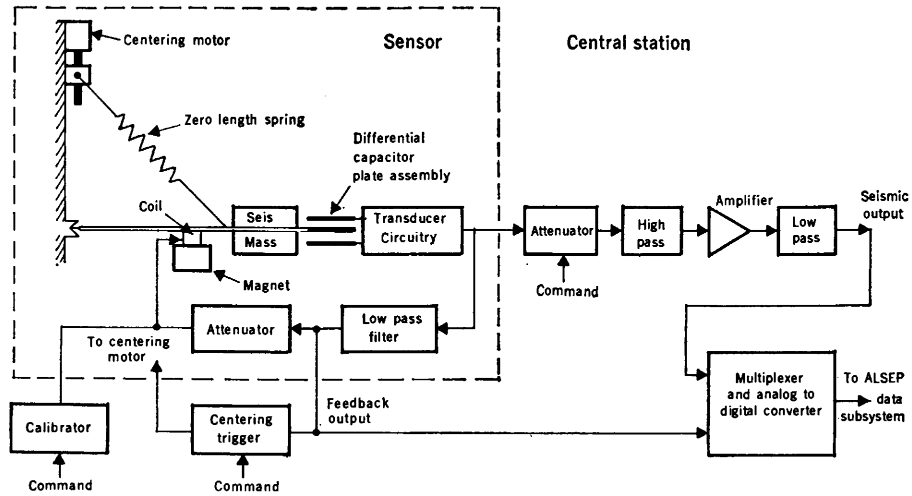


Fig. 7. Schematic diagram of a feedback-controlled seismometer.

Present Knowledge of the Lunar Interior

Our present knowledge of the internal structure and state of stress of the moon is comparable to what was known of the interior of the earth just after the turn of the century when efficient seismographs were first developed. The lunar mass and mean radius are well determined at 7.353×10^{25} grams and 1738 kilometers, respectively (*5*). The moon is, therefore, slightly more than one-quarter the size of the earth. The mean density of the moon calculated from these values is 3.34 grams per cubic centimeter, which is comparable to the density of rocks just below the crust of the earth and is considerably less than the mean density of the earth (5.54 grams per cubic centimeter). This density also corresponds to the class of meteorites called chondrites. Gravitational acceleration at the mean lunar surface can thus be computed to be 162 centimeters per second, or about one-sixth of the earth's surface gravity. Maximum pressures within the moon are equal to those existing in the earth at the relatively shallow depth of 150 kilometers.

Information on the density of the surface rocks on the moon is provided by the Surveyor alpha-scattering experiments (*6*). The results show that the composition of material at the landing sites (Surveyors 5, 6, and 7) resembles that of terrestrial basalts. The density of terrestrial basalts is commonly near 3.0 grams per cubic centimeter. Thus, with due awareness of the fact that the sample is very small, we can infer that the density of lunar surface material beneath the rubble layer is near 3.0 grams per cubic centimeter.

Information on the rate of change of the density of lunar material with depth can be obtained from measurements of lunar moments of inertia. Solomon and Toksoz (*7*) point out that the moment of inertia is very sensitive to changes in surface density and largely insensitive to changes in central density. Values of the principal moments of inertia, based upon recent Lunar Orbiter data (*8*), are

$$A/MR^2 = B/MR^2 = 0.396 \pm 0.009$$

and

$$C/MR^2 = 0.38 \pm 0.13$$

where A is the least and C is the greatest moment of inertia, and M and R are the lunar mass and mean radius, respectively. The apparent reversal in magnitudes between A and C is explained by the much larger uncertainty in C. The actual value of C must, of course, be larger than A. The calculated moment of inertia for a moon with uniform chemical composition, but with allowance for increased density with depth owing to compression, as given by Nakamura and Latham (*9*), is 0.401 ± 0.001. This model also includes the effects of thermal expansion for a central temperature of 1500°K. Thus, it appears that only a moderate increase in density with depth can exist on the moon. The increase may be caused by self-compression alone, although, as Levin (*10*) points out, the effect of thermal expansion on material deep within the moon may more than offset compression by pres-

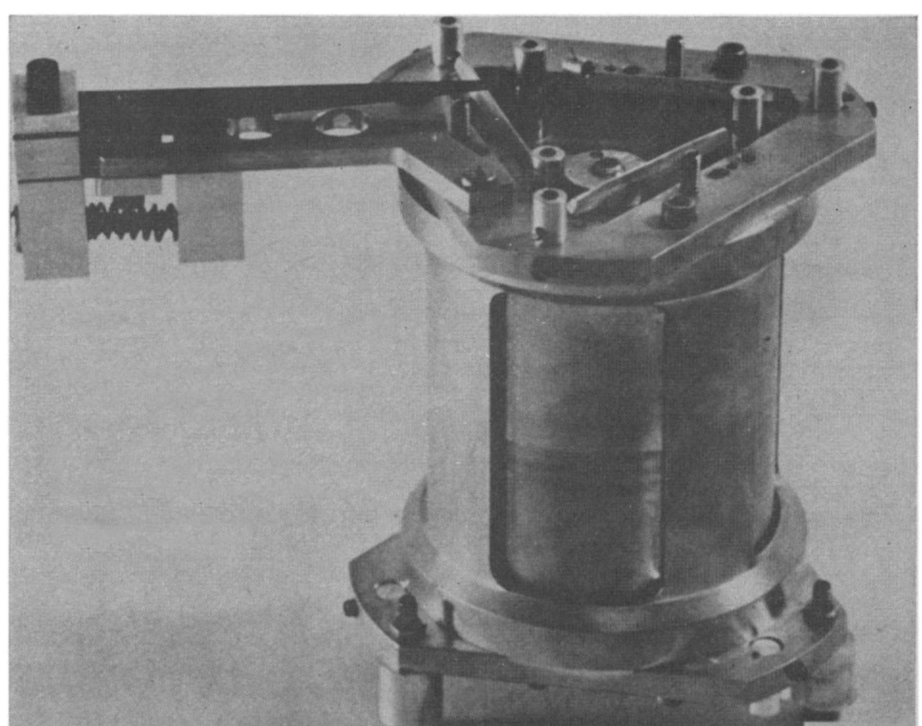


Fig. 8. Photograph of the high-frequency seismometer.

sure. On this basis, Nakamura and Latham (*9*) conclude that a slight increase in the mean molecular weight of lunar material with depth probably does exist. From considerations of moments of inertia and assuming a value of 3.1 grams per cubic centimeter for the density of material at the lunar surface, Solomon and Toksoz (*7*) obtain a density of 3.7 grams per cubic centimeter at the center of the moon if the moon is considered to be chemically homogeneous, and 3.5 grams per cubic centimeter for the central density of a differentiated moon.

The present range of uncertainty in values for the lunar moments of inertia would allow a heavy core (7.5 grams per cubic centimeter) with radius up to 0.3 of the lunar radius, or an outer crust up to 260 kilometers thick if composed of rocks of the density of basalt. It is clear from the difference in average density that the moon must contain substantially smaller percentages of heavy elements, for example, iron, than the earth and may have an average composition similar to the upper mantle of the earth.

A question of great importance to lunar seismic studies is the present thermal state of the interior of the moon. Estimates of lunar temperature distribution have been given by MacDonald (*11*), Levin (*10*), Maeva (*12*), Phinney and Anderson (*13*), and many others. These studies indicate that the moon may or may not be molten in its deep interior, depending upon what values are chosen for concentrations of radioactive materials and for thermal conductivities.

Anderson and Phinney (*14*) conclude that a molten core is likely unless (i) extensive volcanism in the past removed excess heat or (ii) the formation of the moon occurred much later than the formation of the earth so that radiogenic heating cannot have proceeded within the moon to the extent required to produce melting.

It seems clear that at least some of the features on the lunar surface indicate that extensive melting has occurred at some time in its history. It can be argued, however, that all such melting occurred as a direct result of impact phenomena. The question of whether a substantial molten core is present should be settled very soon after moonquake data are obtained.

Sources of Seismic Energy

Expected sources of natural seismic activity on the moon can be broadly divided between internal sources and external sources. Internal sources relate ultimately to internal thermal energy. The mechanism for the release of seismic energy is sudden rupture or change of volume of material within the moon. Internal sources of heat need not be concentrated near the center of the moon but may be distributed throughout the moon in small pockets.

Calculations on the rate of energy release caused by internal heating have been made for various lunar models by MacDonald (*11*). His results indicate that the number of moonquakes produced by internal heating will be at least as high as that on the earth. Press *et al.* (*15*) have estimated that if natural seismic activity on the moon is the same as that on the earth, an average of between 10 and 100 moonquakes per month will occur with magnitudes sufficiently large to produce detectable seismic waves over the entire surface of the moon. If this assumption is correct, a great many smaller moonquakes would be expected to occur that would be detected by a seismometer at a given location. Thus, if the moon is as seismically active as the earth, several hundred moonquakes per month may be recorded.

External sources of seismic activity on the moon include thermal stresses produced by rapid temperature variations at the surface, tidal stresses exerted upon the moon principally by the earth and the sun, and meteoroid impacts. Of these, meteoroid impact is likely to be the only important source of activity.

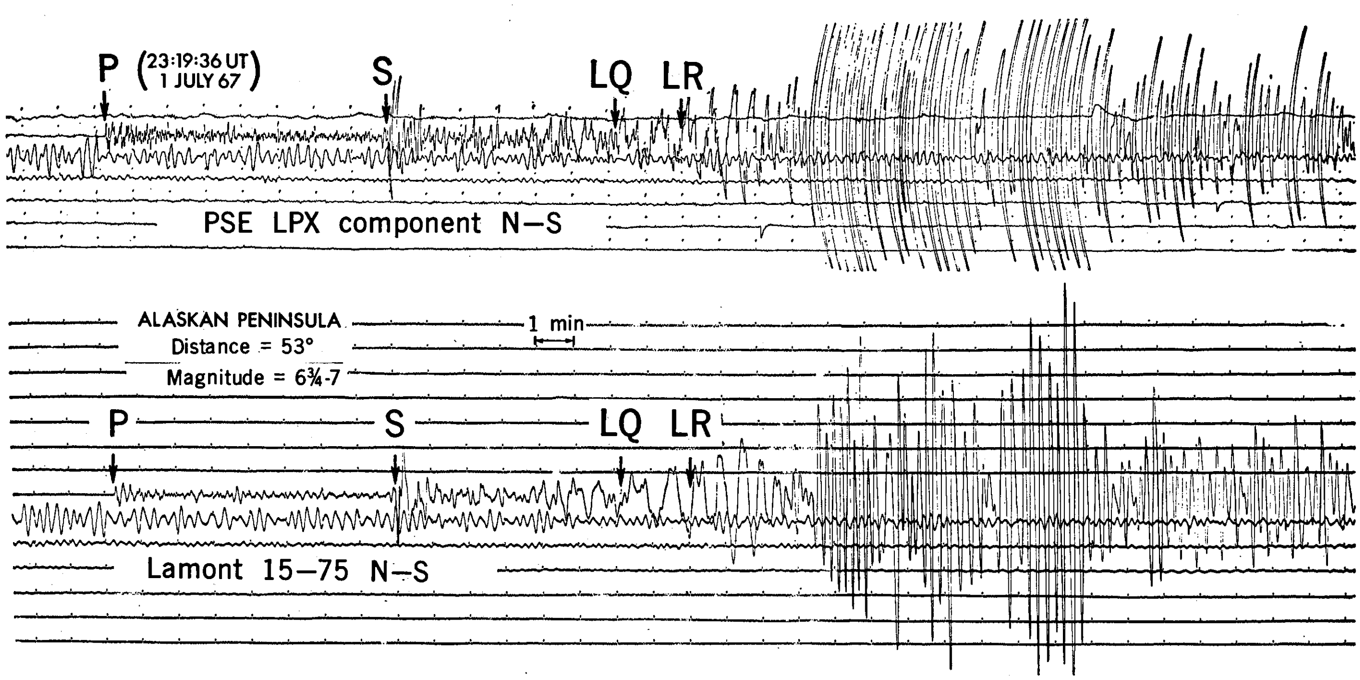
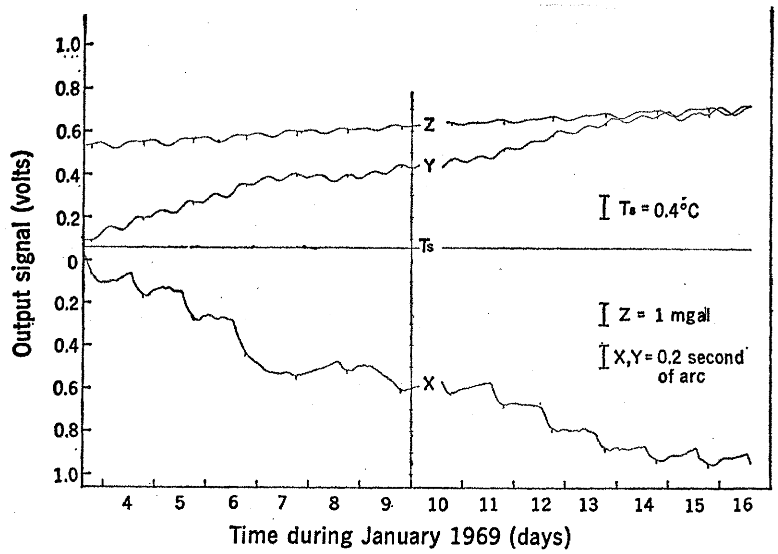


Fig. 9. Record from an earthquake made during testing of the Apollo seismic system (upper section) compared with a recording of the same earthquake from a conventional seismometer-galvanometer combination (lower section). The conventional seismometer system has been in use for many years at the Lamont-Doherty Geological Observatory. Comparisons of this type have shown that the performance of the Apollo seismometer equals or exceeds that of its conventional counterpart in all respects. The major seismic phases (body waves *P* and *S* and surface wave *LQ* and *LR*) are shown in these records. Note that some difference in appearance between the two records is caused by the fact that the Apollo seismometer recorder used in this test has curvilinear pen motion, whereas the photographic recorder used for the conventional seismometer produces a rectilinear trace motion.



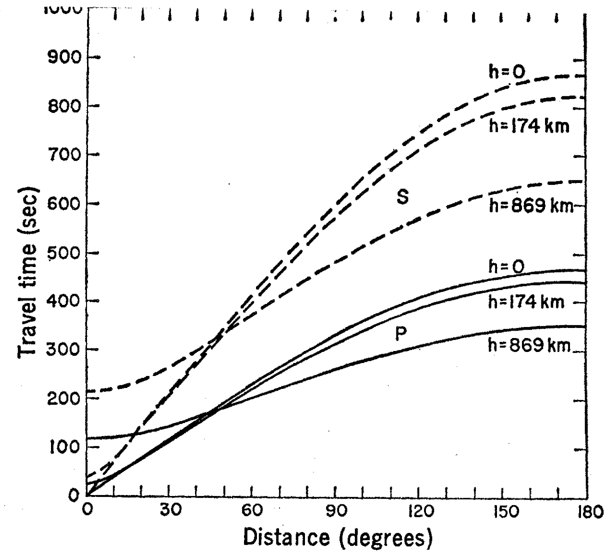


Fig. 10 (left). Sample of tidal data from the Apollo seismic system (feedback outputs); T_s is the output from the instrument temperature sensor; Z is the vertical component; X and Y are the horizontal components. Full-scale output signal is 5 volts. Fig. 11 (right). Theoretical curves of lunar travel time for body waves (compressional waves P and shear waves S for various depths (h) of the moonquake source beneath the surface. The model is for a self-compressed, chemically homogeneous moon with a temperature at its center of 1500°C. [From (*20*), Fig. 4]

Seismicity associated with the impact of meteoroids on the lunar surface has been the subject of several studies. Press *et al.* (*15*) and Laster and Press (*16*) assume an equivalence between seismic signals from underground nuclear explosives of a given yield and those from impacts of meteoroids having kinetic energy equal to the yield.

McGarr *et al.* (*17*) determined the coupling between impacting projectiles and seismic waves from model experiments in which small plastic pellets were fired at hypervelocities into various target materials. The resulting seismic waves were recorded by an array of miniature accelerometers. These studies indicate that the predicted number of impacts which may be recorded depends strongly on the properties of the assumed surface material, but recordings from 100 or 200 impacts may reasonably be expected during the course of 1 year. Most of these will strike within 10 to 20 kilometers of the seismometer, but a few detectable impacts may occur at ranges exceeding 1000 kilometers. Thus, the seismic experiment may also be considered a measurement of meteoroid flux with the entire surface of the moon serving as a collector.

Based upon their experimental data, McGarr *et al.* (*17*) predict that the impact of the spent Saturn S-IVB stage of the Apollo booster, with a mass of 16,000 kilograms and an impact velocity of 2.6 kilometers per second, should be detectable by the lunar seismometer at a distance of at least 200 kilometers. This would serve as an extremely important source of data since the time and location of the impact could be precisely determined. The possibility of doing this experiment on Apollo 12 or 13 is presently under review.

Analysis Techniques and Expected Results

Seismological science is not an end in itself; rather, it is the principal tool for revealing physical state, stress state, and composition of lunar or planetary interiors. Analysis of samples returned from the moon may determine lunar composition to depths as great as 50 kilometers because material has probably been excavated to this depth by meteoroid impacts. These samples can be studied in the laboratory under varying conditions of pressure and temperature, and the elastic properties to the very center of the moon can thus be predicted. Seismic data can then be used to check this extrapolation. Differences between the results of these two methods can be interpreted in terms of temperature, physical state (partial melting), and compositional variations.

It is likely that our progress in lunar seismology will parallel that of terrestrial seismology with the restriction that we will initially be limited to one seismic station. Nevertheless, we will immediately discover the nature of the background seismic noise on the lunar surface. Such noise is likely to be much less important on the moon than on the earth owing to the absence of an atmosphere, oceans, and cultural activities on the lunar surface. It would not be surprising, in fact, if instrumental noise is seen as background. Records of the footsteps of the astronaut as he moves away from the seismometer will give an immediate indication of the effectiveness of the installation.

We will next turn our attention to the seismicity of the moon, that is, the numbers of moonquakes and the magnitude of energy released. It is reasonable to assume that the principal seismic phases [compressional (P) and shear (S) waves, and surface waves] will be identifiable on lunar seismic records. The records will also be searched for phases which might be recognized as reflections from prominent boundaries such as a molten core or the base of a crustal layer. Measured time intervals between arrivals of the observed seismic phases can then be used to obtain rough distances to the source (epicenter) of the moonquakes by comparison with theoretical curves of travel time such as those given in Fig. 11. The direction to the source can be estimated by measurement of the relative amplitudes of the P-wave arrival (or Rayleigh or Love waves) on the two horizontal-component seismometers and the phase relations between the horizontal and vertical components of the surface waves.

Thus, from a single triaxial seismometer, we can obtain crude locations for moonquakes (or impacts) and can determine whether these events occur primarily in a few active regions, as is the case for the earth, or if they are randomly distributed. The correlation between such active regions, if extant, and observed lunar surface features such as rills, highlands, or mare boundaries will be of great interest, as will the question of whether moonquakes tend to occur at shallow depths or deep within the lunar interior or are uniformly distributed in depth. After these questions are answered, it will be possible (i) to elucidate the tectonic mechanisms associated with major crustal features and to determine whether they are still active and (ii) to indicate the present state of thermoelastic stress, and hence the present thermal regime within the moon.

Theoretical seismic velocity curves for the moon have been given by Press *et al.* (*15*), Bolt (*18*), Kovach (*19*), Nakamura and Latham (*20*), and Derr (*21*). A feature common to most of the theoretical lunar models is a broad low-velocity zone. It is possible, in fact, that the seismic velocities within the moon decrease monotonically from the surface to the center. This will make interpretation of data on body waves more difficult, since the decrease of velocity with depth causes the seismic waves to be bent downward with the possibility of quite complicated ray paths.

A second problem which may complicate interpretation of data on body waves on the moon is the possibility of lateral inhomogeneity. It is entirely possible that the moon consists of a multitude of large blocks embedded within a matrix of finer material. The simplifying assumptions involving homogeneous layers usually made in seismic interpretations would then have to be abandoned. However, the dis-

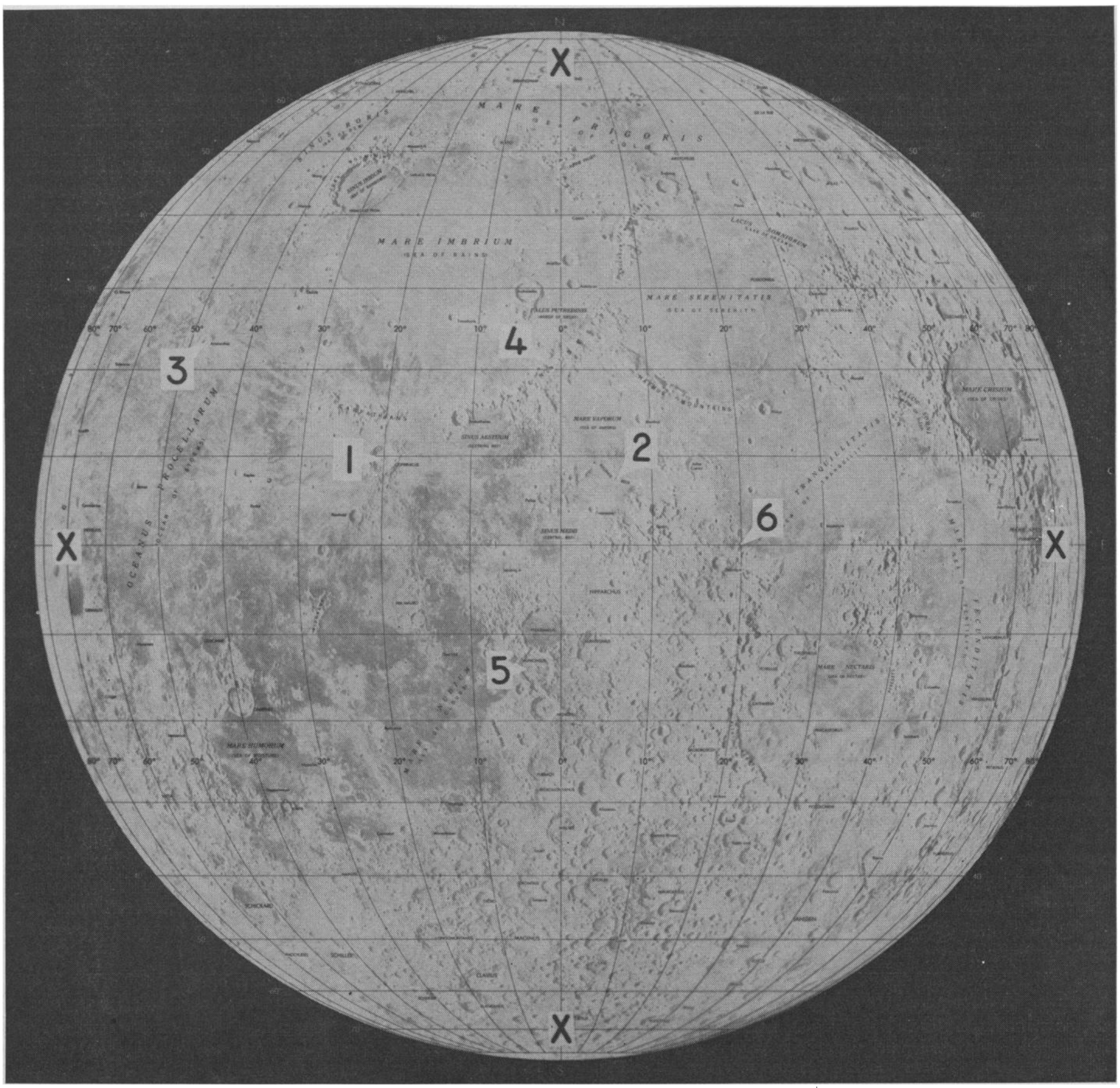


Fig. 12. Suggested locations for an array of lunar seismic stations: (1) floor of Copernicus; (2) Hyginus rill; (3) Aristarchus, floor or surrounding area; (4) Apennine Mountain front; (5) Alphonsus, floor or surrounding area; (6) Sabine and Ritter craters. The X marks indicate sites selected for unmanned lunar landings to complete a lunar seismic net. [From (*25*), Fig. 3]

covery of such lateral variations would have a profound effect on our notions of lunar evolution and the viscoelastic properties of the interior.

When data from an adequate number of quakes have been obtained, we will next want to determine curves of travel time from our measurements and the structure of the moon from such measurements. For body waves, we can begin this process if we are given artificial sources of energy, such as the impact of the Saturn S-IVB mentioned above, or if several stations are in operation simultaneously. For surface waves, we can begin to construct curves of travel time with data from a single station, if we are able to record waves which have traversed the major as well as the minor arc to the seismometer or those which have traversed the circumference of the moon more than once and have been recorded on successive passes. Then the distance traveled is simply the circumference of the moon, and the time interval between successive passes is easily measured on the record. To study these, or their counterparts for the body waves, we must hope for a few strong shocks.

Surface waves will be useful in the study of the moon's outer layers. In terrestrial studies, comparison between theoretical and observed dispersion of surface waves has been the primary method for the determination of the thickness of the earth's crust under continents and oceans. The same methods will be applicable to the study of highlands and mare regions on the moon. Data from studies of the dispersion of surface waves can be combined with orbital geodetic data or direct gravity measurements to determine if the tendency for elevated regions to achieve hydrostatic equilibrium with underlying material (isostasy) exists on the moon as it does on the earth. The relative amplitude between surface waves and body waves can also be used to give at least a qualitative measurement of the depth of the source.

As mentioned above, the low-period seismometers have sufficient sensitivity at ultralow frequencies to make possible the detection of changes in gravity and tilting of the lunar surface associated with lunar tides. Calculations of theoretical tides for assumed lunar models have been made by Sutton *et al.* (*22*), Harrison (*23*), Bolt (*24*), and Derr (*21*). Comparison between theoretical and observed tides will place additional constraints on parameters of the possible lunar model.

Sensitivity at ultralow frequencies also makes possible the measurement of the free oscillations of the moon as a whole if they are excited by large moonquakes or meteoroid collisions. Free oscillations should not be considered physically distinct from surface waves; they simply represent the low-frequency (or long-wavelength) end of the surface wave spectrum. For a homogeneous moon, the fundamental mode of oscillation (spheroidal mode) would have a period of about 14.6 minutes; the presence of a heavy central core decreases the period, and a reversal in density increases it (*19*). Thus, measurements of free oscillations will provide an additional means for differentiating between possible lunar models.

Free oscillation and tidal measurements are particularly attractive in that interpretation of data requires knowledge of the spectrum of the signal from a single station only. Neither the source location nor the precise time of the event is needed. Events large enough to produce observable free oscillations of the earth are rare, but on the smaller (or possibly quieter) moon, they may be more numerous.

We have discussed possibilities for data interpretation based on data from a single station; however, according to present plans, we can expect to have at least two and possibly three stations in operation simultaneously within the next year. This will allow unique locations of epicenters, construction of accurate curves of travel time, and determination of focal depth under favorable circumstances.

Interpretation of lunar seismic data will be a challenging problem. The question of the uniqueness of the interpretation which occurs on the earth will be even more severe on the moon because of the reduced amount of data and, hence, the fewer constraints on possible models. On the other hand, the total pressure variation on the moon is much less than on the earth, and there is some evidence that the compositional variation with depth may be less. A number of elegant techniques are being readied to invert lunar geophysical data. For example, Monte Carlo methods will yield a family of models consistent with the data; hopefully, these models will converge as the data increase to a band of models so narrow that meaningful interpretations will be possible. These models will yield seismic equations of state which, as mentioned above, can be compared with experimental equations of state for returned lunar samples as measured in the laboratory under conditions of pressure and temperature which duplicate the entire range of possibilities at all depths within the moon. In this way seismic data will provide the first indication of the variation of temperature, physical state, and composition within an extraterrestrial body.

Implications of a Dead Moon

A question which often arises in discussions of lunar seismology is, "What if there are no moonquakes?" If this is the case, it should be noted that the use of seismic methods for the determination of lunar structure would not be impossible, although results may come more slowly as a result of having fewer sources. There would still be meteoroid impacts and later, perhaps, artificial sources to provide seismic energy. If the moon is devoid of seismic activity from internal sources, this leads to the conclusion that there are no significant internal energy sources and that the moon could not have a partially molten mantle or a molten core.

As discussed above, the absence of a hot interior, in turn, indicates either (i) that excess heat has been removed in the past by extensive volcanism; or (ii) that the proportion of radioactive materials in the moon is much lower than that for the earth or for meteoroids; or (iii) that the moon was formed much later than the earth, and radiogenic heating has not had time to produce important effects. Thus, the complete absence of internal seismic activity in the moon will be of considerable significance relative to the thermal history of the moon.

Next Step—Post-Apollo

Before discussing plans for future lunar seismic experiments, let us first consider what we may reasonably expect to have achieved by the end of the Apollo program which presently includes four surface landings. It is probably not too optimistic to assume that, by some combination of Apollo missions, three ALSEP's will have operated successfully on the moon by the end of the present Apollo program, each ALSEP having operated for the better part of a year. If we are fortunate, the operational lifetimes of these

ALSEP's may overlap to give us simultaneous recording from several seismic stations. We will have recorded several hundred meteoroid impacts, most of them within 10 to 20 kilometers of the seismometer, moonquakes, if extant, and seismic energy from at least one artificial source—probably the impact of a spent Saturn IV stage or that of a missile launched from earth. Thus, we will have answered the most basic question: At this stage of its evolution does the moon continue to release energy in the form of moonquakes, or is the moon a cold, dead body devoid of seismic activity from internal sources?

By the end of the present Apollo program, we will very likely have achieved at least crude curves of travel time for body waves and surface waves and the beginning of a seismicity map for the moon. From these data, the velocities of compressional and shear waves within the moon will be determined with a precision dependent upon the number, type, and distribution of sources, both natural and artificial. Our state of knowledge going into the post-Apollo program will be greatly improved if we have recorded at least one large seismic event from which detectable surface waves travel several times around the moon. As mentioned above, velocities of surface waves can be calculated from such data independent of the time of origin or location of the source.

If we are fortunate, we will have recorded one or two seismic events of such great magnitude that they will have excited the free modes of vibration of the moon. The precise measurement of the periods and damping of free oscillations will provide an independent determination of the elastic properties of the entire moon. Interpretation of data on free oscillations does not require knowledge of the time or location of the source; thus, data from a single station are sufficient.

During the post-Apollo era, seismologists will want primarily to achieve a wider distribution of detectors in order to map the seismically active belts in greater detail and to study the mechanisms of energy release. Goals of lesser importance will be (i) to lower the minimum detectable ground motion of an individual seismometer, and (ii) to improve the performance of long-period seismometer systems at the ultralong-period end of the spectrum for recording surface waves from moonquakes, free oscillations of the moon, and lunar tides.

A possible distribution of stations previously suggested by Latham (*25*) is shown in Fig. 12. In this site selection, features which might be expected to be seismically active and which are of interest in other respects are emphasized. The central peak of the crater Alphonsus (site 5), for example, is the feature from which Kozyrev and Ezerski in 1958 (*26*) observed the escape of a small quantity of carbon gas. The Apennine Mountains (site 4), which are nearly as high as the Himalayas, represent the effects of tremendous crustal stresses and might reasonably be expected to be a region of high seismicity. Craters Sabine and Ritter (site 6), which were photographed by Ranger 8 and Lunar Orbiter 4, have been interpreted as possible volcanic calderas. Aristarchus (site 3) is the brightest crater on the moon and has associated with it the most intense infrared anomaly thus far detected. Hyginus rill (site 2) is the most prominent of the rills—features which appear to be large grabens in some cases. The X marks (Fig. 12) indicate sites at which seismometers might be emplaced by unmanned lunar landing missions to augment the seismic network. Emplacement of additional stations from roving or orbiting vehicles has not been assumed since these systems are not presently planned.

Of course, many other sites could be chosen. The deployment scheme shown in Fig. 12 is based on the assumption that there will be six additional manned lunar landings after the four missions presently planned. This network of stations is certainly sparse compared to the situation on the earth. It is estimated that seismographs presently operate at between 600 and 1000 locations on the land surfaces of the earth. If we take into account the fact that the surface area of the moon is 1/16 that of the earth, 30 seismograph locations are needed on the near side of the moon to achieve the same number per unit surface area as on the earth. Even the present number of earth stations is considered too small for detailed studies. However, the network shown in Fig. 12 would provide enough coverage to permit us to answer many fundamental questions about the moon.

The need for the simultaneous operation of the stations in this network places heavy emphasis upon the need for long-lived systems. These stations must be designed to operate for at least 2 or 3 years with 5 years as a design goal.

Conclusions

We stand at the threshold of the most exciting experiment in seismology. We can for the first time hope to probe beneath the surface of an extraterrestrial body into its deep interior, to listen to its "heartbeat." Although only a single lunar seismological station can be established by the Apollo 11 mission, the capabilities of that station are so great that it may well prove to be superior to the majority of existing terrestrial stations. Within the next decade it should be possible to greatly expand our understanding of the solar system by use of seismic techniques, not only on the moon, but also on Mars and possibly Venus. The tools and the methods are available. We have only to resolve to do the job.

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